

Seismic anisotropy, mantle fabric, and the magmatic evolution of Precambrian southern Africa

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ABSTRACT

The observed seismic anisotropy of the southern African mantle from both shear-wave splitting and surface wave observations provides important constraints on modes of mantle deformation beneath this ancient continent. We find that the mantle anisotropy beneath southern Africa is dominated by deformational events in Archean times occurring within the lithosphere, rather than present-day processes in the sublithospheric mantle. Consequently, the distribution and magnitude of anisotropy provide valuable data to constrain the mantle's role in the tectonic evolution of this region.

The pattern of mantle anisotropy reveals several noteworthy characteristics. First, mantle anisotropy is closely associated with the Great Dyke of the Zimbabwe Craton, with values of the splitting fast polarization direction, ϕ , parallel to the Dyke. This correspondence with the Great Dyke is likely not due to the present-day Dyke structure but instead is most probably due to the emplacement of the Dyke parallel to pre-existing mantle fabric within the Zimbabwe craton. This deformation thus predates dike emplacement and is no younger than Neo-Archean in age. Second, there is a spatially continuous arc of mantle anisotropy extending from the western Kaapvaal Craton to the northeastern Kaapvaal and Limpopo Belt. All along the arc, ϕ is subparallel to the trend of the arc. Given the crust/mantle chronology associated with these regions, the anisotropy likely represents deformation that occurred at ~2.9 to ~2.6 Ga during collisional accretion of both the western Kimberley and northern Pietersburg blocks onto the seismically isotropic eastern shield of the Kaapvaal, with accretion on the northern ramparts of the Kaapvaal ultimately culminating in the Neo-Archean Limpopo orogen. The anisotropy-inferred arc of deformation reveals diverse zones of both strong and weak coupling between the crust and mantle, as measured by the coherence between mantle deformation and geologically-inferred surface deformation. In particular, there is high coherence between surface and mantle deformation at the southwestern and northeastern ends of the arc, which implies strong crust-mantle coupling in these regions. Conversely, apparent decoupling exists in the northwestern portion of the arc, where northeast to southwest trending anisotropy cuts across north to south trending structures, such as the surface outcrop and aeromagnetic expressions of the Kraaipan Greenstone Belts. Independent seismic evidence from seismic reflection profiling supports the conclusions that these north-south-trending crustal features are superficial and confined to the upper crust.

We present evidence that the mantle fabric producing seismic anisotropy constitutes fossil structure in the mantle that is subsequently reactivated, much like the more commonly acknowledged reactivation of crustal structures. In particular, we argue that Neo-Archean collisional orogenesis imparted a mechanical anisotropy to the mantle that controlled the subsequent magmatic history of cratonic southern Africa. We furthermore suggest that four major Precambrian magmatic events: the Great Dyke, the

Ventersdorp, Bushveld, and the Soutpansberg, all represent extensional failure along planes oriented parallel to the local splitting fast polarization direction. Each of these events is interpreted to be a collisional rift, similar to the Baikal rift of northern Eurasia, where the stress field associated with collision produces extension and rifting for orientations at a small angle to the direction of the collision. Precise crustal geochronology associates both Ventersdorp and Great Dyke magmatism with the earliest and latest phases of the Limpopo collision, respectively. Similarly, the Bushveld magmatic event is temporally linked to the ~ 2.0 Ga reactivation of Neo-Archean structures in the Limpopo and surrounding areas by the Magondi Orogen, and the Soutpansberg is related to the ~1.9 Ga Kheis Orogen. Since the timing of these basaltic intrusions is controlled by temporal variations in lithospheric stress associated with orogenesis, it implies either that the melting process is genetically related to the evolution of the far-field collision, or that there was a semi-permanent reservoir of basaltic magma residing in the sublithospheric mantle during the ~1 billion-year time period spanned by these magmatic events. The existence of an extensive magma reservoir would argue for elevated temperatures just beneath the lithosphere during this time.

Splitting delay times, δt , a measure of the magnitude of anisotropy, reveal geologically controlled variations in the strength of anisotropy. In particular, the Meso-Archean Kaapvaal shield, the area that was not exposed to ~2.9 Ga and later deformational events, is effectively isotropic. We observe two areas where the anisotropic/isotropic transition is relatively sharp. The north-south boundary appears to coincide with the east-west trending Thabazimbi-Murchison Lineament. In the west, the boundary has been observed in the vicinity of Kimberley, South Africa, near the Colesberg Magnetic Lineament. The Eastern Shield has been relatively devoid of the kind of rifting and magmatic events seen elsewhere in cratonic southern Africa since the Meso-Archean, suggesting that the Eastern Shield lithosphere is mechanically stronger than surrounding areas. This relative strength difference may in part be due to the absence of the mechanical anisotropy inferred for the surrounding areas.

Introduction

What distinguishes Archean cratons from other kinds of continental lithosphere? This single question has occupied Earth Scientists for decades, and continues to do so. While we have not yet obtained a satisfactory answer to this question, there is broad agreement that it is the mantle, and the mantle's interaction with the crust, that is fundamentally different. For example, the tectosphere model for the creation of Archean cratons (Jordan, 1975; 1988), posits the existence of a buoyant, highly refractory basalt-depleted mantle root that serves to stabilize, strengthen, and thus increase the survivability of ancient lithosphere. Southern Africa is particularly valuable in providing access to this mantle component. The abundance of kimberlite nodules from the mantle and crust has had a dramatic impact on our understanding of cratonic southern Africa. The Kaapvaal Project, and in particular the deployment of the Southern African Seismic Experiment, has allowed for an unprecedented examination of the lithospheric mantle beneath an Archean craton.

Despite the mantle's central importance to understanding cratons, the tectonic history of the region has, of necessity, been primarily derived from the geology of crustal rocks. Given the importance of the mantle, it is essential to obtain the tectonic history of the cratonic mantle. We seek answers to several questions. For example, we see terrane boundaries at the surface. Are these lithospheric boundaries or are they superficial, upper crustal features? More generally, do the crust and mantle evolve and deform as a coherent unit, in a way idealized by a thin viscous sheet, or do they behave differently, due to compositional and mineralogical differences in rheology. Do they indeed have a distinctly different history? As with the geological study of the crust, we seek fossil indicators preserved in the lithospheric mantle that may provide constraints on this history.

There are various seismic structures that could be used as mantle fossil indicators. One such indicator is the presence of seismic reflectors, denoting zones of high impedance contrast that can be interpreted as changes in either mineralogy or composition. For example the Moho has been mapped beneath southern Africa (Nguuri *et al.*, 2001), and is clearly geologically controlled. Reflectors interpreted as ancient dipping slabs have been reported using data from the Slave Craton (Bostock, 1997). We have carefully examined the discontinuity structure of the upper mantle (between the Moho and 410km depth) beneath the Kaapvaal Craton for features similar to those found beneath the Slave craton. A systematic examination of these data, however, revealed a very simple upper mantle, with no evidence for equivalent reflectors (Gao *et al.*, 2002).

Another particularly valuable way to access the tectonic history of the mantle is through the study of mantle seismic anisotropy. The anisotropy of the lithospheric mantle reflects strain-induced fabric that, coupled with available geochronology, provides constraints on the structural-geological history of mantle deformation. One may then proceed, like a structural geologist whose field area is the mantle, by attempting to infer the underlying process that led to the deformation, and then comparing this structure to structures in the crust. Also, as with the crust, this fabric is not only an indicator of past deformation, but it may also modulate subsequent deformation.

As we will see, the mantle fabric derived from seismic anisotropy is dominated by the now-well-documented ~2.9 Ga and later Neo-Archean collision between the Eastern Kaapvaal Shield and surrounding cratonic regions to the west and north. The pattern of deformation is much simpler than that inferred from crustal rocks, and reveals regions of coherent deformation between crust and mantle, and regions that do not display this correspondence. We show that this

mantle fabric produced oriented mechanical weakness (mechanical anisotropy) in the mantle, and that this fabric was reactivated during four subsequent large-scale rifting and magmatic events: the Great Dyke, Ventersdorp, Bushveld, and Soutpansberg.

Seismic velocities are a function of propagation and polarization direction and this dependence is referred to as seismic anisotropy. Seismic anisotropy in the mantle is most likely due to lattice-preferred orientation (LPO) in mantle minerals, primarily olivine, which is produced by finite strain in the mantle (i.e., Silver and Chan, 1988; Ribe, 1989; Karato and Wu, 1993). Extensive knowledge concerning the relationship between LPO and strain permits inference of the mantle deformation field using observations of seismic anisotropy. One particularly robust manifestation of seismic anisotropy is referred to as shear-wave splitting. With shear wave splitting, a shear wave propagating through an anisotropic medium is split into two quasi-shear waves with orthogonal polarizations and differing velocities. The orientation and magnitude of the anisotropic layer can then be characterized by a fast polarization direction, ϕ , and delay time, δt . The phenomenon is analogous to the optical birefringence exhibited by certain minerals such as calcite or olivine. For vertically-propagating shear waves, in a simple anisotropic system, and for a shear flow characterized by a horizontal flow line (lineation direction), ϕ will be oriented parallel to the flow line, and δt will be proportional to the product of the thickness and intrinsic anisotropy of the layer.

There are two sources of deformation that are expected to dominate contributions to mantle anisotropy: lithospheric deformation and asthenospheric deformation. The simplest form of the first type is referred to as vertically coherent deformation of the lithosphere (Silver, 1996), which is idealized by the deformation of a thin viscous sheet (England and McKenzie, 1982), and characterized by a vertical velocity gradient approaching zero (Silver, 1996). In this case the crust and mantle deform coherently. This kind of anisotropy is most efficiently generated by transpressional deformation, producing a vertical shear plane and horizontal flow line. Asthenospheric deformation may be idealized by simple asthenospheric flow, representing the horizontal shear flow between two high-viscosity layers: the lithosphere above and subasthenospheric mantle below (e.g., Vinnik *et al.*, 1995). It is characterized by a horizontal shear plane and horizontal flow line. In the case of vertically coherent deformation, the surface strain field can be used as a predictor of the mantle field. In simple asthenospheric flow, the orientation of shear (and hence ϕ) is parallel to the differential velocity vector between the lithosphere above and high-viscosity mantle below (Silver and Holt, 2002).

Examples of each type of contribution, as well as the presence of both, exist throughout the world. An excellent example of vertically coherent deformation is the mantle deformation of the Tibetan Plateau, in

response to the India-Eurasian collision (McNamara *et al.*, 1994; Silver *et al.*, 1999; Holt, 2000). Western North America corresponds to a region that is dominated by simple asthenospheric flow, where the subasthenospheric mantle is moving to the east with a velocity of about 5cm/yr (McNamara *et al.*, 1994; Silver *et al.*, 1999; Holt, 2000). Eastern North America possesses characteristics of both models, with a component of sublithospheric mantle flow around a continental keel, combined with a clear signature of vertically coherent deformation within the lithosphere in some regions (Fouch *et al.*, 2000).

Southern African mantle anisotropy

The Southern African Seismic Experiment (SASE) has permitted the unprecedented sampling of the seismic properties of the lithospheric mantle (e.g., Carlson *et al.*, 1996; Freybourger *et al.*, 2001; James *et al.*, 2001; Nguuri *et al.*, 2001; Silver *et al.*, 2001; Gao *et al.*, 2002; Niu and James, 2002; Saltzer, 2002; Fouch *et al.*, 2004a; b). In the following, we focus on those studies that have analyzed seismic anisotropy.

Shear-wave splitting observations

As observed by Silver *et al.* (2001), mantle anisotropy has been observed throughout the Western Kaapvaal, Zimbabwe Craton, and Limpopo belt but is only weakly present in the eastern Kaapvaal Shield and off-craton to the south and west (Figure 1). The values of ϕ exhibit systematic spatial variations. In the southwestern Kaapvaal they are roughly north-northeast to south-southwest, rotate to northeast-southwest further north, and to nearly east-west in the northeastern part of the craton, including the Limpopo belt. Just north of the Limpopo, there are several stations in the vicinity of the Great Dyke with ϕ values oriented north-northeast to south-southwest. Values of ϕ range from about 0° to 80° (clockwise from north), and δt values for the entire region are small. Where detected, δt is roughly half of the global average of 1.0 s (Silver *et al.*, 2001), and splitting was not detected at ~25% of the stations. An assessment of the crustal contribution (through the analysis of splitting in P-to-S conversions from the Moho), suggests that the contribution to δt is primarily from the mantle, with the crustal component estimated to be about 20% of the total splitting value (Silver *et al.*, 2001).

There are clear regional variations in the size of splitting delay times within cratonic southern Africa. As noted by Silver *et al.* (2001), while there is observable anisotropy in the north and west of the craton, the eastern shield is effectively isotropic. The lateral transition between strong and weak anisotropy is fairly abrupt in some places. For example, there is a clear north-south transition that is approximately coincident with the EW trending Thabazimbi-Murchison Lineament (TML) (Figure 1). In addition, in the central Kaapvaal there is evidence for a sharp transition in fabric strength, based on data from the high-resolution Kimberley

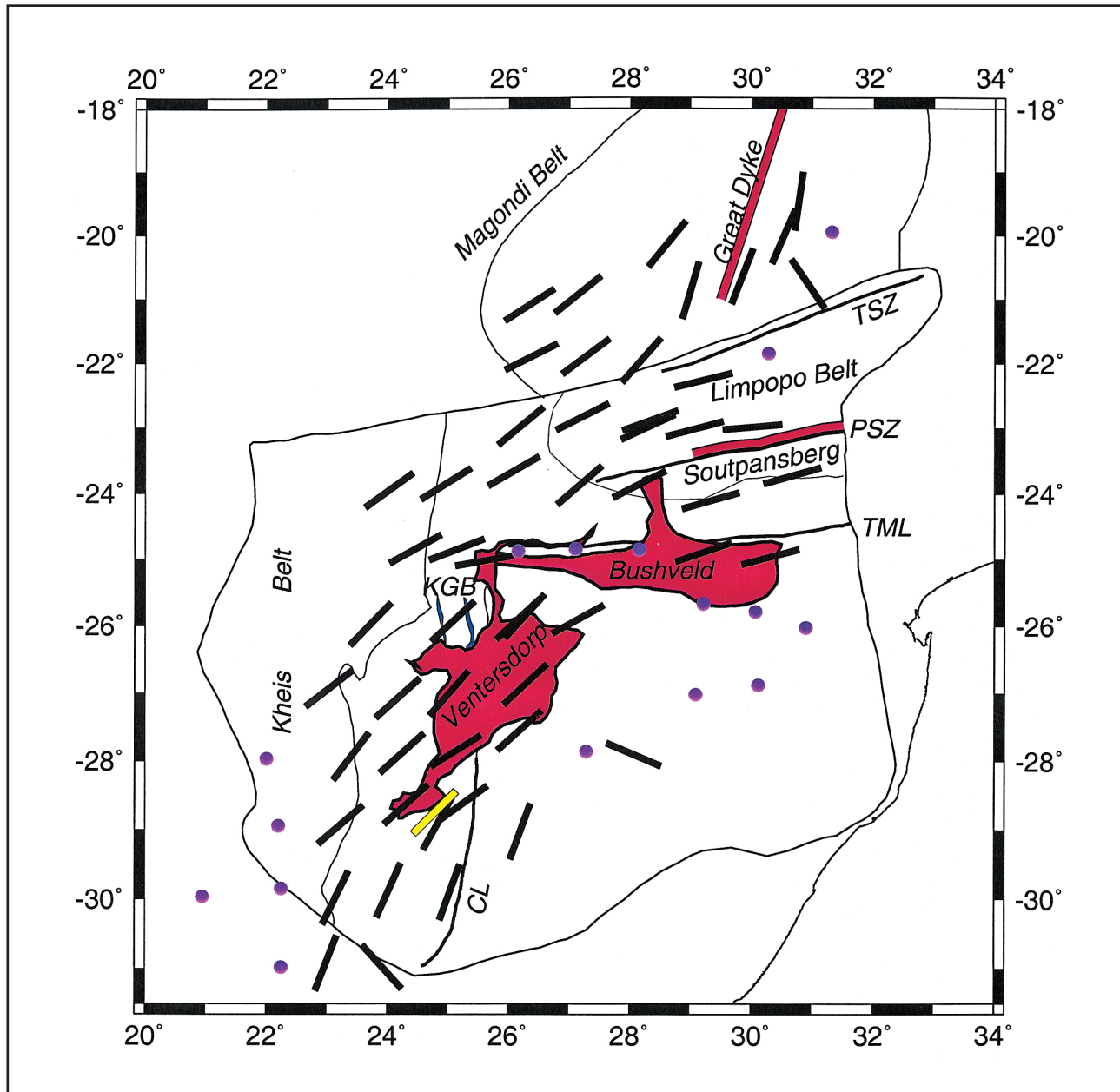


Figure 1. Map showing orientations of shear wave splitting fast polarization directions, ϕ , from the Kaapvaal Seismic Array (black bars) and from the high-resolution Kimberley Seismic Array (yellow bar, array average). Only highest-quality splitting measurements are shown. Purple dots represent measurements with zero or near-zero splitting delay times. Also shown are relevant geological features. Abbreviations: TML: Thabazimbi Murchison Lineament, CL: Colesberg Lineament, KGB: Kraaipan Greenstone Belts, SZ, shear zone. Major rift systems with basaltic magmatism: Great Dyke, Ventersdorp (surface exposures shown), Bushveld and Soutpansberg shown in red. The splitting reveals several interesting features: (1) an arc of values of ϕ from the southwestern Kaapvaal craton to the Limpopo and northern Kaapvaal Craton, with ϕ parallel to the trend of the arc. (2) Values of ϕ that are parallel to the Great Dyke in the Zimbabwe craton. (3) The contrast between relatively weak anisotropy within the eastern Kaapvaal shield, and stronger anisotropy in surrounding areas to the west and north. (after Silver *et al.*, 2001).

Seismic Array (KSA, Fouch *et al.*, 2004b), which consists of 31 stations with an aperture of about 60x40km. As shown in Figure 2, there is a remarkably systematic variation in SKS splitting delay time across the array, from northwest to southeast, decreasing from about 0.8 to 0.2 s over only 60 kilometers. Because of the small aperture of the array, these variations in splitting parameters must result from differences in lithospheric mantle fabric.

Surface wave anisotropy

A study of radial and azimuthal anisotropy in surface waves reveals the following characteristics. First, the magnitudes of anisotropy inferred from the surface and body wave data sets are compatible (e.g., Freybourger *et al.*, 2001; Saltzer, 2002). An integral of the surface-wave model yields a surface-wave-derived estimate of the splitting delay time of 0.6s, which is essentially the same as the average splitting delay time. The average

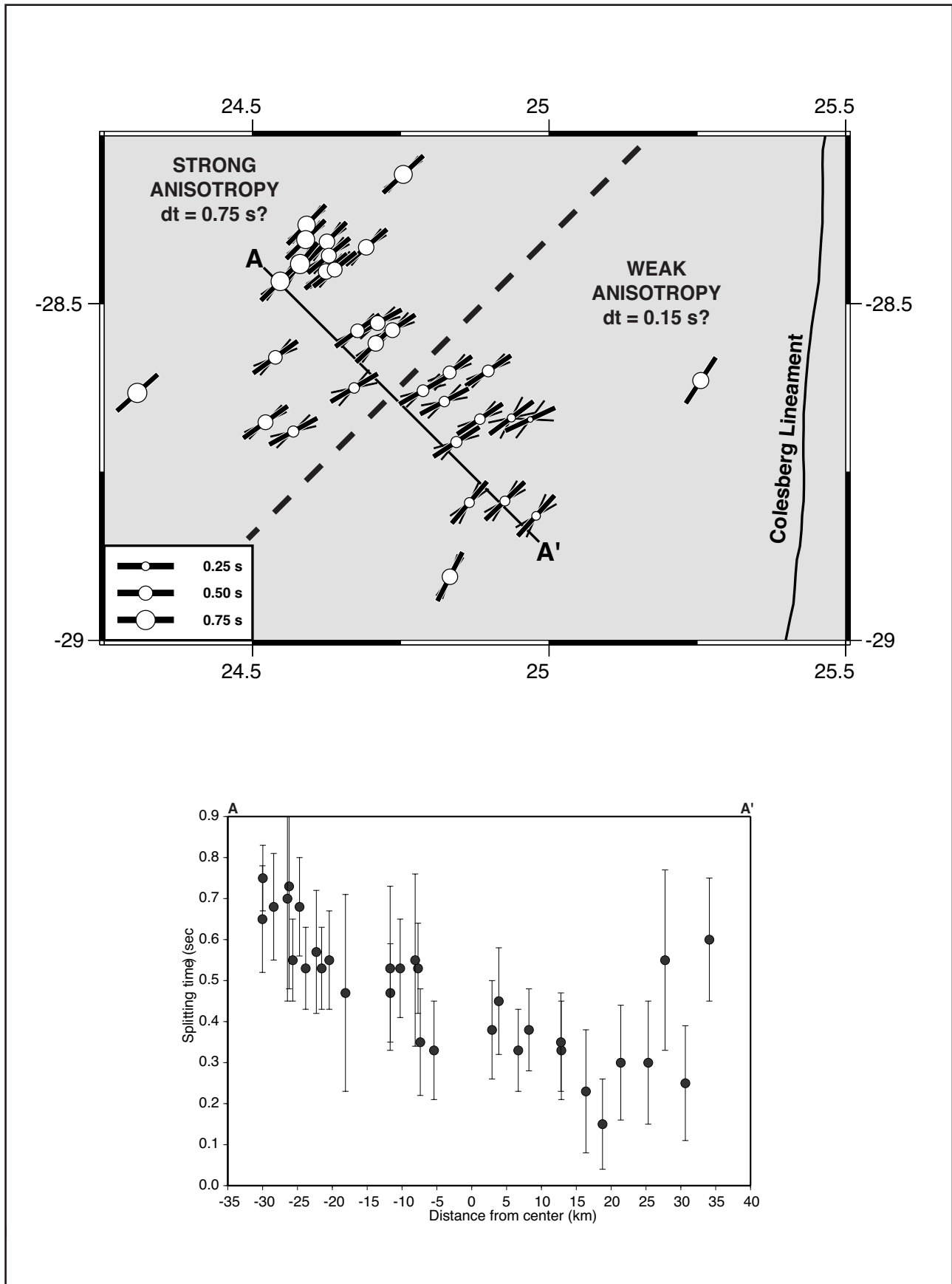


Figure 2. Shear wave splitting results for the Kimberley region Kimberley Seismic Array from Fouch *et al.* (2004b). (Top:) Azimuth of black bars denote fast polarization direction and standard deviation; size of open circles is scaled to splitting time; splitting time scale shown in bottom left corner. Splitting delay times show a smooth increase from southeast to northwest, and range from 0.15 s to 0.73 s (bottom). Proposed boundary of anisotropic/isotropic domain delineated by dashed line (see text for details).

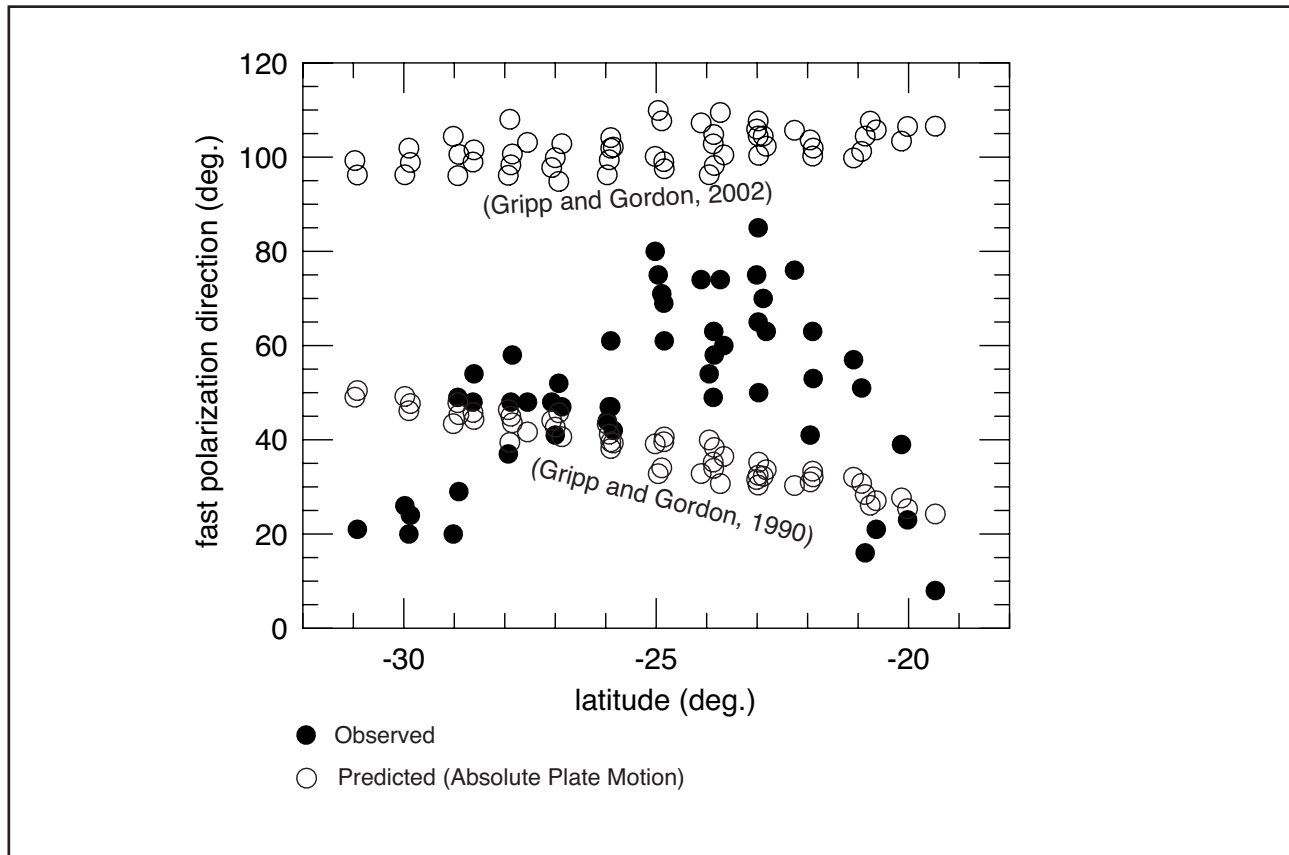


Figure 3. Comparison of Kaapvaal Seismic Array shear wave splitting results with absolute plate motion (after Silver *et al.*, 2001). Observed values (clockwise from north) of ϕ (filled circles) for stations shown in Figure 1 (highest quality), excluding three outliers [sa10, sa28, and sa77 with (latitude, ϕ) = (-30.97, -42), (-27.90, -67) and (-20.76, -34) respectively], and predicted values of ϕ from absolute plate motion (APM) model HS2-NUVEL1A (lower open circles) and newer HS3-NUVEL1A (upper open circles). APM-predicted values also show systematic latitudinal dependence, but which is distinctly different from the data. Note that both plate-motion models provide a poor fit to the data.

depth extent of the anisotropy inferred from surface waves indicates that it is primarily of mantle origin and is localized within the lithosphere to depths no greater than 220km. There was no detectable azimuthal anisotropy when averaged over the entire data set. This observation is consistent with the large range of splitting fast polarization directions found throughout the regions that have likely been averaged out in the surface-wave analysis.

Physical process responsible for mantle anisotropy

We next seek to determine the physical process (or processes) responsible for the mantle anisotropy. In order to use the anisotropy to study ancient lithospheric deformation, we first need to determine whether this is indeed the dominant source of anisotropy. Toward this goal, we first evaluate the alternative model that the anisotropy is due to simple asthenospheric flow, which would correspond to present-day deformation within the asthenosphere that is the result of the motion of the African plate over the deeper mantle. We can test the simple asthenospheric flow model by making a prediction of ϕ using models of both plate motion and the velocity of the

subasthenospheric mantle. For example, we can assume that the subasthenospheric mantle is fixed in a hotspot reference frame. Figure 3 shows the observed values of ϕ as a function of latitude, as well as the predictions from two models for this case: NUVEL-HS2-1A (Gripp and Gordon, 1990), and a newer model NUVEL-HS3-1A (Gripp and Gordon, 2002). The earlier model makes predictions that are consistent with some of the observations (where the two trends cross, at latitudes of -28° and -20°), but it fails to account for the overall trend in ϕ . The more recent plate-motion model is an even poorer fit to the observations. It is important to point out that the absolute velocity of Africa is not well constrained, since Africa plate velocities are very low; these two models are not significantly different from each other, nor are they significantly different from a velocity of zero for the African plate (Gripp and Gordon, 2002). However, virtually any plate-motion model would yield a linear trend in Figure 1, and thus fit the data poorly. Another possibility is that the subasthenospheric mantle beneath southern Africa is not fixed in a hotspot reference frame, but in motion. Indeed analysis of splitting for oceanic stations surrounding southern Africa reveals the influence of a large-scale lower-mantle upwelling (Behn *et al.*, 2003). However, the inferred

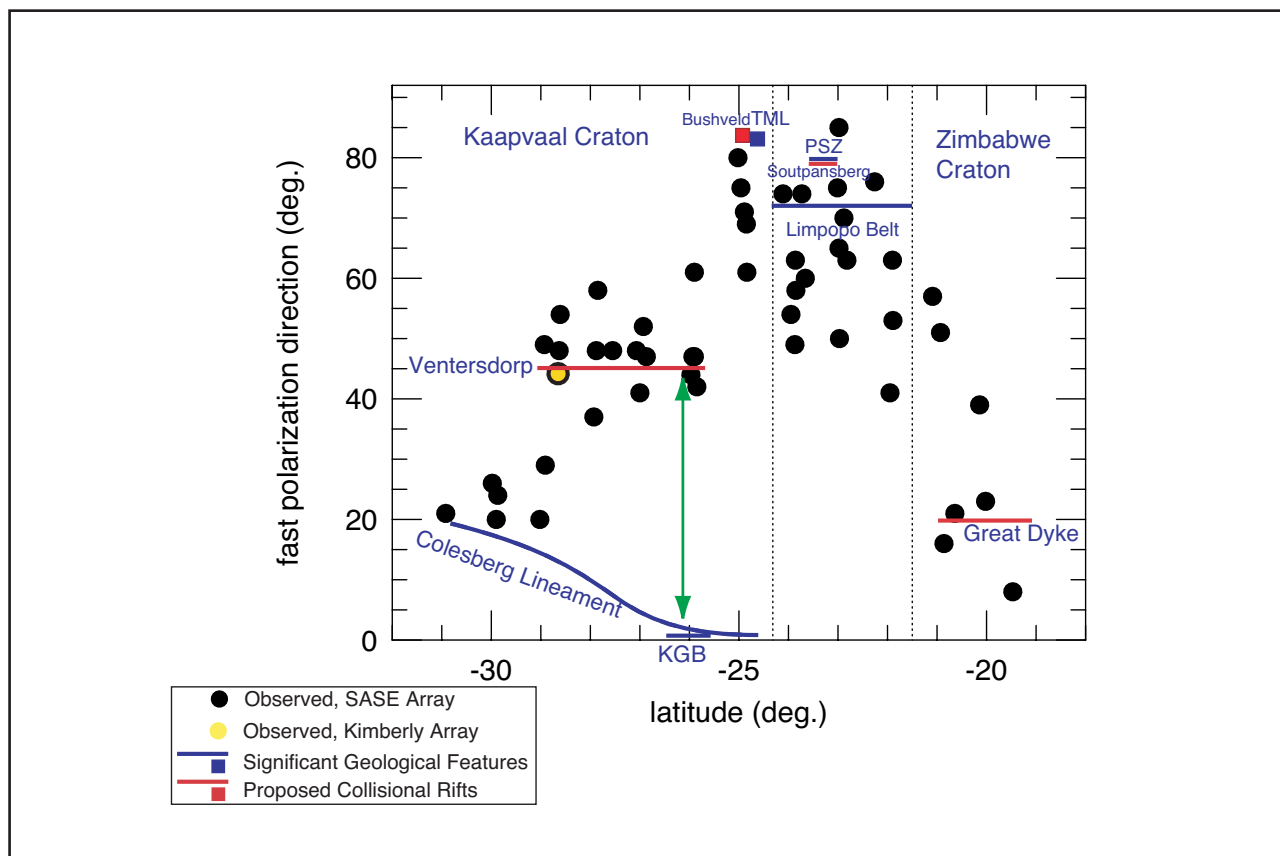


Figure 4. Comparison of Kaapvaal Seismic Array shear wave splitting results with primary surface geologic features of cratonic southern Africa. Approximate latitudes and orientations of prominent geologic features and boundaries, given by horizontal lines [Colesberg Lineament, Kraaipan Greenstone Belts (KGB), Limpopo Belt, Great Dyke, Ventersdorp], filled boxes (Thabazimbi-Murchison Lineament, Bushveld, Soutpansberg). Vertical dashed lines give approximate latitude range of the Limpopo Belt. Note that data exhibit strong latitudinal dependence that closely tracks these geologic features in most, but not all, areas. A notable exception is in northwestern Kaapvaal craton (Latitude range 25°S to 27°S), where splitting fast polarization directions are at a high angle to KGB and CML (vertical green arrow).

flow-field is approximately uniform beneath southern Africa and predicts northeast-southwest values of ϕ (similar to that predicted by the NUVEL-HS2-1A) and is also a poor fit to the data. These comparisons thus suggest that simple asthenospheric flow is not the dominant source of mantle anisotropy beneath southern Africa.

Another possible model is that both a lithospheric and deeper asthenospheric component exist in the region. This model geometry would lead to two anisotropic layers, which have a predictable effect on the back-azimuthal variations in splitting (Silver and Savage, 1994; Rumpker and Silver, 1998; Saltzer *et al.*, 2000). We have in fact searched for two anisotropic layers but have found no evidence for such structure (Silver *et al.*, 2001). Thus, we conclude that the splitting observations argue for a lithospheric source that is unrelated to present-day tectonics. We note that a lithospheric source for the anisotropy is corroborated by regional surface wave studies as noted above (Freybourger *et al.*, 2001; Saltzer, 2002), high-resolution measurements in the Kimberley region of the Kaapvaal Craton (Fouch *et al.*, 2004b), and an extensive examination of mantle xenolith petrofabrics, clearly

documenting that the Kaapvaal Craton lithospheric mantle is seismically anisotropic (Ben-Ismaïl *et al.*, 2001).

Characterizing the mantle lithospheric deformation field and its relation to surface geology

The pattern of anisotropy mentioned above implies a pattern of ancient mantle lithospheric deformation beneath southern Africa. There are three striking characteristics of mantle deformation across the region that may be inferred from the seismic anisotropy (Figure 1). First, on the largest scale, we observe a spatially continuous arc of mantle deformation that extends from the southwestern Kaapvaal to the northeastern Kaapvaal Craton, Limpopo belt and southwestern Zimbabwe Craton. Along this region, ϕ is oriented subparallel to the arc. This pattern of deformation is typically observed in modern convergent margin settings dominated by transpressional deformation, such as Tibet (McNamara *et al.*, 1994; Sandvol *et al.*, 1997; Holt, 2000). The inner edges of this arc of deformation are clearly delimited in the southwest by the Colesberg Lineament (CL) and in the northeast by

the Thabazimbi-Murchison Lineament (TML) on the Kaapvaal Craton, while the outer edge of the arc to the north is demarcated by the Triangle shear zone of the Limpopo Belt (Figure 1). The arc is primarily associated with ~2.9 to ~2.6 Ga convergent margin processes dated in both the crust and mantle (de Wit *et al.*, 1993; Carlson *et al.*, 2000; Kroner *et al.*, 2000; Poujol *et al.*, 2002; Schmitz, 2002; Schmitz *et al.*, 2003), and thus likely records an Archean mantle deformational event. Second, the older ~3.1 to ~3.7 Ga eastern shield of the Kaapvaal Craton is characterized by weak to negligible anisotropy. Finally, there is a distinctly different mantle deformation field, oriented approximately north-northeast–south-southwest in northeastern Zimbabwe, that is geographically associated with and oriented parallel to the Great Dyke. As we will see, Great-Dyke-related fabric is also likely Archean in age (Figure 1).

One of the important issues in understanding lithospheric evolution in the Archean is whether the crust and mantle components of the lithosphere deform coherently. The extensive seismic coverage of cratonic southern Africa from the SASE has allowed us to examine crust-mantle coherence in unprecedented detail. In many parts of southern Africa, coherence of mantle anisotropy with large-scale crustal structures is observed, (Figures 1 and 4), with splitting directions paralleling the north-northeast trending Colesberg Magnetic Lineament in the southwestern Kaapvaal, and the east-northeast trending structures (Triangle and Palala Shear Zones and the Thabazimbi-Murchison Lineament) in the northeastern Kaapvaal and Limpopo belt, which represent the ends of this large-scale deformational arc. However, in the middle region, namely the northwestern Kaapvaal craton, crustal deformation (as inferred from limited geologic fabric data) has a distinctly different orientation, compared to deformation inferred from mantle anisotropy. In this area, the arc of mantle deformation exhibits northeast-southwest trending fabric (Figure 1). In contrast, the surface geology reveals north-south trending structures, in particular the Kraaipan Greenstone Belts, which at face value would suggest significant crust-mantle decoupling in this area. Although it is admittedly difficult to identify the dominant crustal fabric of the region, given the paucity of basement exposure and the significant overprint of Ventersdorp extensional tectonics in the region, such an inference is supported by a recent reexamination of seismic reflection profiles, concluding that these north-south trending features are probably superficial and limited to the upper crust (de Wit and Tinker, 2004). The structural complexity of this region is unsurprising, given that it was the syntaxis of synchronous ca. 2.9 Ga deformation and collisional accretion along both the northern and western ramparts of the Kaapvaal shield (de Wit *et al.*, 1992; Poujol *et al.*, 1996; Kroner *et al.*, 2000; Schmitz, 2002; Schmitz *et al.*, 2003).

One of the most intriguing features of the splitting data set is the anisotropy and inferred mantle

deformation associated with the Great Dyke. This locale is particularly valuable to our analysis because the close correspondence between the Great Dyke orientation and ϕ ensures that the mantle anisotropy is unambiguously reflecting Archean structures in the mantle. As argued by Silver *et al.* (2001), it is highly unlikely that the Great Dyke structure itself would give rise to the anisotropy. If one attempts to explain, for example, the shear-wave splitting patterns with shape preferred orientation (SPO) geometries, due to the compositional difference between the Dyke and surrounding rocks in the mantle, there is simply insufficient seismic velocity contrast to generate the observed anisotropy. Dikes could only generate such a splitting signal if there existed a significant melt fraction, which is unlikely for the Great Dyke at present.

A much more probable explanation is that the Great Dyke was extruded through pre-existing lithospheric mantle fabric that possessed this particular orientation. This places a minimum date for the fabric of ~2.58 Ga, the emplacement age of the Great Dyke (Mukasa *et al.*, 1998, Wingate, 2000). The relationship between this mantle fabric and the local geologic history of the surface rocks of the Zimbabwe Craton, however, is unclear. There have been two primary crust-forming, deformation-inducing episodes in the Zimbabwe craton at ~3.5 Ga and ~2.9 Ga. There is no obvious relationship between this mantle fabric and the ~2.9 Ga event. The ~3.5 Ga event may have been widespread at one time (Horstwood *et al.*, 1999). The largest remnants of this earlier episode are found in two locales with nearly identical ages: the Tokwe segment and the Midlands, just east and west of the Great Dyke respectively. However, the dominant orientation of crustal fabric in these remnants is unclear. The mantle fabric thus appears unrelated to the ~2.9 Ga event, and may or may not be related to the ~3.5 Ga event. In either case, it appears that there has been significant decoupling between the upper crust and upper mantle from the time of this fabric formation to the subsequent emplacement of the Great Dyke. Crust-mantle decoupling has also been inferred from the geology, based on the distribution pattern of primary accretionary shear zones (Dirks and Jelsma, 2003).

Pre-existing mantle fabric and magmatic evolution

There is a well-known axiom in geology that the style of crustal deformation is often controlled by pre-existing structures, or in short, “old faults never die”. Indeed, a crustal fault can be reactivated many times within the context of differing stress regimes. The TML, active from ~2.9 Ga to the Mesozoic, is one of the world’s best examples (Good and de Wit, 1997). Application of the same concept to pre-existing structures in the lithospheric mantle is less commonly discussed, but no less important. Indeed, if deformation involves the entire lithosphere, and if the mantle represents the strongest portion of the lithosphere, then such structures could in

Table 1.

Rift Event	Ventersdorp	Great Dyke	Bushveld	Soutpansberg
Age (Ga) [ref]	2.71 [1]	2.57 [2,3]	2.06 [4,5]	1.88 [6,7]
Orogen (fabric-forming)	Kimberley-Pietersburg*	?	Pietersburg-Limpopo	Pietersburg-Limpopo
Age (Ga) [ref]	2.93-2.88 [8-11]	?	2.93-2.67 [8-14]	2.93-2.67 [8-14]
Rift Orientation (wrt ϕ **)	0	0	15	0
Orogen (rift-forming)	Limpopo(CZ-SMZ)	Limpopo(CZ-NMZ)	Magondi	Kheis
Age (Ga) [ref]	>2.72-2.67 [12-14]	>2.67-2.52 [13-17]	>2.04-2.00 [18-20]	1.93-1.75 [21]
Rift Orientation (wrt σ **)	45	40	45	10

* We informally refer to the orogenies occurring at ca 2.9 Ga on the western and northern margins of the Kaapvaal shield the Kimberley and Pietersburg Orogens, respectively. σ denotes direction of horizontal most compressive stress direction for orogen.

** wrt: with respect to.

References: [1] Armstrong *et al.* (1991); [2] Mukasa *et al.* (1998); [3] Wingate (2000); [4] Walraven and Hattingh, 1993; [5] Buick *et al.* (2001); [6] Bumby *et al.* (2002); [7] Hanson *et al.* (this volume); [8] de Wit *et al.* (1992); [9] Kroner *et al.* (2000); [10] Poujol *et al.* (2002); [11] Schmitz [2002]; [12] Barton *et al.* (1992); [13] Kroner *et al.* (1999); [14] Kreissig *et al.* (2001); [15] Mkweli *et al.* (1995); [16] McCourt and Armstrong (1998); [17] Frei *et al.* (1999); [18] Majaule *et al.* (2001); [19] Mapeo *et al.* (2001); [20] McCourt *et al.* (2001); [21] Cornell *et al.* (1998)

fact play the dominant role in determining the style of lithospheric deformation. It has long been noted that continental (lithospheric) breakup often occurs where collisions have previously occurred. The opening of the Atlantic basin in North America near the suture of the closing of the Iapetus Ocean is a classic example. Recently, it has been argued that the deformation-induced lattice preferred orientation (LPO) that produces mantle seismic anisotropy, also produces a mechanical anisotropy. The study of mechanical anisotropy in olivine aggregates (olivine is the weakest of the mantle minerals and thus controls strength) suggests that shear failure is most probable along the plane (010). For transpressional deformation, the (010) plane is expected to be oriented vertically with strike parallel to the trend of the collisional belt and to the splitting fast polarization direction, ϕ . This hypothesis thus predicts that former collisional zones would be weak in shear and tension for vertical planes oriented parallel to the trend of the collisional belt (Tommasi and Vauchez, 2001). We consequently extend the old geological axiom to the mantle: "old fabric never dies". In the case of Southern Africa, we propose that the mantle fabric, acquired through collisional deformation in the Neo-Archean, has played a major role in controlling the subsequent extensional and magmatic evolution of the region.

While the Great Dyke represents the clearest case of a dike swarm passing through lithosphere with preexisting mantle fabric, there are three other cases where the same phenomenon appears to be taking place, namely the Ventersdorp Rift and flood basalts (2.715 to 2.709 Ga, Armstrong *et al.*, 1991), the Bushveld Complex (2.059-2.054 Ga, Walraven and Hattingh, 1993; Buick *et al.*, 2001), and the Soutpansberg Trough (~1.9 Ga, Bumby *et al.*, 2002) (Figures 1 and 4). In the case of the Ventersdorp, recent analysis of seismic reflection profiles strongly suggests that the dominant orientation of Ventersdorp rifts is northeast to southwest (Tinker *et al.*, 2002; de Wit and Tinker, 2004, with the notable exception of east-west trending faults just south of the

TML). Throughout this zone, the average value of ϕ is about 45°, parallel to these faults. Similarly, the Bushveld is localized along the east-west trending TML (du Plessis and Walraven, 1990), and its location corresponds to a latitude where there is an abrupt change in ϕ from 45° to a value closer to east-west. The Sibasa basalts were erupted into the Soutpansberg Trough, a half-graben bounded on the south by reactivation of the Palala Shear Zone as a north-dipping normal fault (Bumby *et al.*, 2002). The axis of the trough is aligned not only with the Palala Shear Zone but also with the east-west trend of mantle anisotropy in the Central Zone of the Limpopo Belt. It thus appears that all four of these magmatic events have passed through and exploited preexisting fabric in the lithospheric mantle.

Collisional rifts

Not only are the orientations of the dike intrusions controlled by mantle fabric created by previous deformational events, but it also appears that in all four cases the magmatic episodes themselves are closely related to contemporaneous deformational events (Table 1). The Neo-Archean Ventersdorp rift is temporally linked with compressional deformation and crustal thickening in the Central and Southern Marginal Zones of the Limpopo Orogen (Barton *et al.*, 1992; Kroner *et al.*, 1999; Kreissig *et al.*, 2001). Similarly, the Great Dyke was emplaced during the exhumation of the Northern Marginal Zone of the Limpopo Orogen over the Zimbabwe Craton (Mkweli *et al.*, 1995; McCourt and Armstrong, 1998; Frei *et al.*, 1999). In the Paleoproterozoic, emplacement of the Bushveld-Molopo Farms Complex was nearly synchronous with the Magondi Orogeny along the Norath-west margin of the Zimbabwe-Kaapvaal Cratons (Majaule *et al.*, 2001; Mapeo *et al.*, 2001; McCourt *et al.*, 2001). Significant dextral transpressional reactivation of the Triangle and Palala Shear Zones of the Limpopo Belt are also recorded at this time (Kamber *et al.*, 1995; Schaller *et al.*, 1999; Bumby *et al.*, 2001). Subsequent extensional

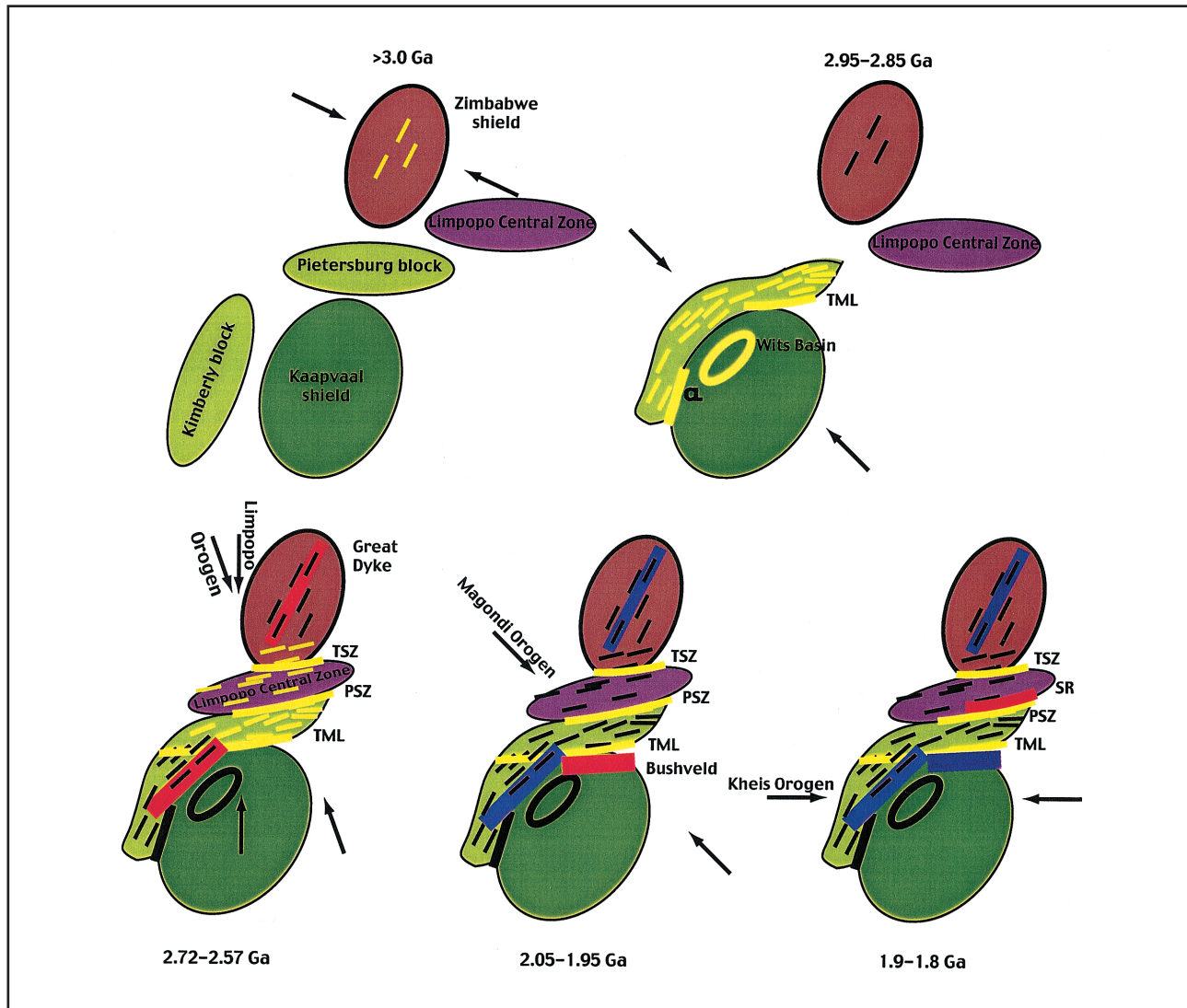


Figure 5. A scenario based on the anisotropic structure of the mantle and its relation to the surface geology of cratonic Southern Africa. Arrows show compression direction of orogen. (i) Formation of the pre-2.9 Ga north-northeast-south-southwest mantle fabric in Zimbabwe Craton and establishment of the isotropic structure of the eastern shield of the Kaapvaal Craton. (ii) Development of north-northeast-south-southwest to east-northeast-west-southwest arc of lithospheric mantle fabric by ca. 2.9 Ga collisional accretion of terranes along the northern and western margins of the Kaapvaal Craton. (iii) Continued eastwest to east-northeast-west-southwest mantle fabric development during and within the ca. 2.7 Ga Limpopo Orogen and its marginal zones within the Kaapvaal and Zimbabwe Cratons. Also during this time, emplacement of the Great Dyke and Ventersdorp supergroup. These are both collisional rifts responding to far-field stresses caused by the latest (north-northwest-south-southeast directed compression) and earliest (north-south directed compression) phases of the Limpopo Orogen respectively, with rift orientations controlled by mantle fabric produced ca. 2.9Ga. (iv) Reactivation of Limpopo structures by Magondi Orogen and the creation of a collisional rift that produced the Bushveld Intrusion. Intrusion orientation (east-west) controlled by preexisting mantle fabric. (v) Finally, creation of the collisional Soutpansberg Rift (SR) by the Kheis Orogen at 1.88 Ga.

reactivation of the Palala Shear Zone and formation of the Soutpansberg Trough coincided with compressional deformation in the Kheis Belt along the western Kaapvaal Craton (Cornell *et al.*, 1998; Bumby *et al.*, 2002).

A plausible explanation for this temporal correspondence is that they are all collisional rifts or impactogens, similar to the Rhinegraben and Baikal rifts (Sengor *et al.*, 1978). These rifts form at a high angle to collisional belts and are due to the far-field stresses associated with the collision. Such rifting is possible for any vertical plane orientation for which the collision

produces tensional normal stresses. This corresponds to orientations that are within 45° of the collision direction. The correspondence between rift orientation and mantle anisotropy arises because the actual rift orientation is locally controlled by the pre-existing mantle fabric.

Thus a collisional rift is expected to form at a high angle to the collisional belt, and be synchronous with or precede the magmatic episode. Indeed Burke *et al.* (1985) and Wilson (1990) have made such a proposal for the Great Dyke and Ventersdorp respectively. Regarding rift orientation, in all four cases the angle between belt and rift is less than or equal to 45° , and consistent with

this mechanism (Table 1). Regarding the timing, there are precise ages both for the orogenic events, in the form of the age of high temperature metamorphism, and for the onset of rifting. For the rifts, these are Great Dyke: ~2.58 Ga (Wingate, 2000; Mukasa *et al.*, 1998), Ventersdorp: ~2.72 Ga (Armstrong *et al.*, 1991), Bushveld: ~2.06 to ~2.05 Ga (Walraven and Hattingh, 1993; Buick *et al.*, 2001) and Soutpansberg: ~1.88 Ga (Bumby *et al.*, 2002; Hanson *et al.*, 2004). The corresponding ages of orogenic high-temperature metamorphism are ~2.58 to ~2.56 Ga for the CMZ (Kroner *et al.*, 1999), ~2.69 to ~2.70 Ga for the SMZ, ~2.03 Ga for the Magondi Orogen, and ~1.93 Ga for the Kheis (Cornell *et al.*, 1998, Table 1).

While these two sets of ages are very close, we note that in 2 of the 4 cases, the Ventersdorp and Bushveld, the orogenic age is actually younger than the rifting age by 20 to 30 Ma. At first glance this would appear to be inconsistent with the collisional-rift hypothesis. Indeed, Tinker *et al.* (2002) made this particular point in arguing that the Ventersdorp is not a collisional rift (and proposing an alternative model). It is, however, important to point out that the rift ages and orogenic ages are fundamentally different. While the rifting ages most likely represent the true onset of rifting, the orogenic ages, being high temperature metamorphic ages, represent either the middle or the end of the orogeny, assuming the metamorphism is due primarily to crustal thickening (rather than magmatic advection). This means that the metamorphic age may be younger than the onset age by 20 to 30 Ma (England and Thompson, 1984; Huerta *et al.*, 1998), which is approximately the observed difference between the rifting and orogenic ages. Accounting for this source of bias, we conclude that, given the presently available data, the onset ages of rifting and orogenies are consistent with the collisional-rift hypothesis. This is a testable hypothesis, in that a more accurate measure of the orogen onset age should yield an age that is equal to or greater than the rift age. If the ages are indeed indistinguishable, it suggests that the rifting begins at the onset of the orogen for these two cases.

Isotropic strength

As is clear from Figure 5, the tectono-thermal activity in the anisotropic zones to the west and north is in dramatic contrast to the relative tranquility of the Eastern Shield, during this same period of time. This appears to be a contrast in strength, in that the Eastern Shield was subjected to comparable stresses, especially with respect to the Limpopo collision. In addition we note that the distribution of kimberlite eruptions, which we take as another measure of lithospheric strength, is much more prevalent in the west, compared to the east. There may be a multitude of factors that could control lithospheric strength. Lithospheric thickness differences are often invoked as a potential factor, for example. There is presently, however, no evidence that the eastern Shield is systematically thicker than regions to the west

and north by the present-day seismically-inferred lithospheric thickness, presuming that it is an indicator of Precambrian thickness as well (see Fouch *et al.*, 2004a). This strength difference may be attributed to mechanical anisotropy in the mantle. An anisotropic aggregate will have a weak orientation that will always be weaker than the strength of an equivalent isotropic aggregate. We thus speculate that the difference in strength between the eastern shield and elsewhere is in part due to the absence of a mantle fabric in the east.

Discussion and conclusions

A summary of the Precambrian history of southern Africa from the viewpoint of the anisotropic mantle is relatively simple, as is shown schematically in Figure 5 and Table 1. We define five major deformational phases, tied to major orogens that control the history of this craton: (i) an unknown pre-2.9 Ga orogen that imparts a mantle fabric to the Zimbabwe craton, (ii), a collision at ~2.9 Ga along the western and northern boundaries of the Kaapvaal Shield, imparting mantle fabric to the Kimberly and Pietersburg terranes, (iii) The Limpopo orogen, at ~2.6 to ~2.7 Ga, which imparts mantle fabric to the three Limpopo zones, and which produces collisional rifts to the north and south, namely the Great Dyke and Ventersdorp, (iv), the ~2.0 Ga Magondi orogen, both reactivating shear zones in the Limpopo and producing the Bushveld as another collisional rift, and finally (v) the ~1.8 to 1.9 Ga Kheis orogen, which produces the final collisional rift, namely the Soutpansberg trough. In all of these cases, the rifts formed at an orientation that is parallel to preexisting mantle fabric, as inferred from mantle anisotropy.

An important conclusion of the work discussed here is the realization of the dual role of orogens, both in creating strain-induced fabric in the mantle, and in subsequently generating far-field extensional stresses that reactivate this mantle fabric in the form of collisional-rifting events (e.g. Tommasi and Vauchez, 2001). The four collisional rifts discussed here are not superficial shallow crustal features, but instead represent rifts through the entire lithosphere, given both the close correspondence with mantle deformation, and the fact that their basaltic-composition magmatic products represent primary melts of the sublithospheric mantle. As such, these rifts constitute useful constraints on the properties of both the lithosphere and sublithospheric mantle. In particular, it is necessary to account for both the source of these basalts, as well as the survivability of cratonic mantle lithosphere, in the face of these magmatic events.

A significant consequence of a collisional-rift interpretation of these magmatic events is that the timing of rifting and magmatism is controlled entirely by temporal variations in lithospheric stress, rather than, say, the transient ascent of plumes rising from the deep mantle. The apparent ready availability of mantle-derived melts during each of these rifting events raises

two intriguing hypotheses. The first requires a temporal relationship between orogenesis and sublithospheric upper mantle melting, namely, could melting in the sublithospheric upper mantle be primed in some way by the pre-history of plate convergence and subduction beneath the cratonic lithosphere? Alternatively, did there exist a semi-permanent reservoir of partial melt beneath the lithosphere that was tapped whenever lithospheric stresses were favorable to the creation of rifts? If so, the timing of the 4 rifts we have discussed would necessitate the survival of this reservoir for nearly one billion years. Completely evaluating these two alternatives is beyond the scope of this contribution; however we may speculate on mechanisms for establishing super-solidus temperatures in the sublithospheric mantle. Estimates of the potential temperature required for melts to occur in the subcontinental asthenosphere are approximately 1450 to 1550°C (White and McKenzie, 1995), some 200 to 300°C hotter than those today. It is well known that thick continental lithosphere has a thermal blanketing effect on the mantle below which result in higher mantle temperatures (e.g. Guillou and Jaupart, 1995; Grigne and Labrosse, 2001) if the rate of heat generation is greater than the rate of heat loss. More generally, beneath large continents the mode of convection locally behaves more like stagnant-lid convection, rather than plate tectonics. This mode of convection is much less efficient at removing heat, which can lead to a gradually increasing potential temperature (e. g. Sleep, 2000), to the point that the geotherm could cross the solidus of fertile peridotite, while at the same time leaving the refractory depleted lithosphere intact. This molten basalt is then able to rise to the surface whenever collisional rifts are formed in the lithosphere.

While we presently lack the evidence to fully evaluate the mechanisms for melt generation associated with these Archean and Proterozoic rifting events in southern Africa, we believe the highlighted correlations with old mantle fabric, collisional orogenesis and changes in plate stresses reveal a path of future study toward a better understanding of the role of continental lithosphere in upper mantle dynamics. Similarly while further data are necessary to firmly establish the collisional-rift interpretation of Archean and Paleo-Proterozoic tectonomagmatic events, this hypothesis has the advantage of making several testable predictions, from the chronology of orogens, to the petrology of basalt magmas, to the mechanical strength of lithospheric mantle. We look forward to this evaluation in future studies of the southern African lithosphere.

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References

- Armstrong, R.A., Compston, W., Retief, E.A., Williams, I.S. and Welke, H.J. (1991). Zircon ion microprobe studies bearing on the age and evolution of the Witwatersrand triad. *Precambrian Research*, **53**, 243-266.
- Barton Jr., J.M., Doig, R., Smith, B.C., Bohlender, F. and van Reenen, D.D. (1992). Isotopic and REE characteristics of the intrusive charnoenderbite and enderbite geographically associated with the Matok Pluton, Limpopo Belt, southern Africa. *Precambrian Research*, **55**, 451-467.
- Behn, M., Conrad, C. and Silver, P.G. (2004). Detection of Upper Mantle Flow Associated with the African Superplume. *Earth and Planetary Science Letters*, submitted.
- Ben-Ismaïl, W., Barruol, G. and Mainprice, D. (2001). The Kaapvaal craton seismic anisotropy: A petrophysical analysis of upper mantle kimberlite nodules. *Geophysical Research Letters*, **28**, 2497-2500.
- Bostock, M.G. (1997). Anisotropic upper-mantle stratigraphy and architecture of the Slave Craton. *Nature*, **390**, 392-395.
- Buick, I.S., Maas, R. and Gibson, R. (2001). Precise U-Pb titanite age constraints on the emplacement of the Bushveld Complex, South Africa. *Journal of the Geological Society of London*, **158**, 3-6.
- Bumby, A.J., Eriksson, P.G., van der Merwe, R. and Brummer, J.J. (2001). Shear-zone controlled basins in the Blouberg area, Northern Province, South Africa: syn- and post-tectonic sedimentation relating to ca. 2.0 Ga reactivation of the Limpopo Belt. *Journal of African Earth Sciences*, **33**, 445-461.
- Bumby, A.J., Eriksson, P.G., van der Merwe, R. and Steyn, G.L. (2002). A half-graben setting for the Proterozoic Soutpansberg Group (South Africa): evidence from the Blouberg area. *Sedimentary Geology*, **147**, 37-56.
- Burke, K., Kidd, W. and Kusky, T. (1985). Is the Ventersdorp Rift system of southern Africa related to a continental collision between the Kaapvaal and Zimbabwe cratons at 2.64 Ga ago?. *Tectonophysics*, **115**, 1-24.
- Carlson, R.W., Grove, T.L., de Wit, M.J. and Gurney, J.J. (1996) Anatomy of an Archean craton: A program for interdisciplinary studies of the Kaapvaal craton, southern Africa. *Eos Transactions, American Geophysical Union*, **77**, 273-277.
- Carlson, R.W., Pearson, D.G., Boyd, F.R., Shirey, S.B., Irvine, G., Menzies, A.H. and Gurney, J.J. (1999). Re-Os systematics of lithospheric peridotites: implications for lithosphere formation and preservation. In: J. J. Gurney, J. L. Gurney, M. D. Pascoe and S. H. Richardson (Editors), Proceedings of the 7th International Kimberlite Conference, **1**, *Red Roof Design, Cape Town, South Africa*, 99-108.
- Cornell, D.H., Armstrong, R.A. and Walraven, F. (1998). Geochronology of the Proterozoic Hartley Basalt Formation, South Africa: constraints on the Kheis tectogenesis and the Kaapvaal Craton's earliest Wilson Cycle. *Journal of African Earth Sciences*, **26**, 5-27.
- De Wit, M.J., Roering, C., Hart, R.J., Armstrong, R.A., de Ronde, C.E.J., Green, R.W.E., Tredoux, M., Peberdy, E. and Hart, R.A. (1992). Formation of an Archean continent. *Nature*, **357**, 553-562.
- De Wit, M.J., Armstrong, R.A., Kamo, S.L. and Erlank, A.J. (1993). Gold-bearing sediments in the Pietersburg Greenstone Belt: age equivalents of the Witwatersrand Supergroup sediments. *South Africa. Economic Geology*, **88**, 1242-1252.
- De Wit, M.J. and Tinker, J.S. (2004). A composite deep-seismic reflection profile across the central Kaapvaal craton. *South African Journal of Geology*, **107**, 185-205.
- Dirks, P. and Jelsma, H.A. (2002). Crust-mantle decoupling and the growth of the Archean Zimbabwe craton. *Journal of African Earth Sciences*, **34**, 157-166.
- Du Plessis, C.P. and Walraven, F. (1990). The tectonic setting of the Bushveld Complex in Southern Africa, Part 1. Structural deformation and distribution. *Tectonophysics* **179**, 305-319.
- England, P. and McKenzie, D. (1982). A thin viscous sheet model for continental deformation. *Geophysical Journal of the Royal Astronomical Society*, **70**, 295-321.
- England, P.C. and Thompson, A.B. (1984). Pressure-temperature-time paths of regional metamorphism I. Heat transfer during the evolution of regions

- of thickened continental crust. *Journal of Petrology*, **25**, 894-928.
- Fouch, M.J., Fischer, K.M., Parmentier, E.M., Wyssession, M.E. and Clarke, T.J. (2000). Shear wave splitting, continental keels, and patterns of mantle flow. *Journal of Geophysical Research*, **105**, 6255-6275.
- Fouch, M.J., James, D.E., VanDecar, J.C., van der Lee, S., and K.S. Group (2004a). Mantle seismic structure beneath the Kaapvaal and Zimbabwe cratons. *South African Journal of Geology*, **107**, 33-44.
- Fouch, M.J., P.G. Silver, P.G., Bell, D.R., and Lee, J.N. (2003b). Small-scale variations in seismic anisotropy beneath Kimberley, South Africa. *Geophysical Journal International*, 10.1111/j.1365-246X.2004.02234.x.
- Frei, R., Blenkinsop, T.G. and Schonberg, R. (1999). Geochronology of the late Archean Razi and Chilimanzi suites of granites in Zimbabwe: implications for the late Archean tectonics of the Limpopo Belt and Zimbabwe Craton. *South African Journal of Geology*, **102**, 55-63.
- Freybourger, M., Gaherty, J.B., Jordan, T.H. and T.K.S. Group (2001). Structure of the Kaapvaal craton from surface waves. *Geophysical Research Letters*, **28**, 2489-2492.
- Gao, S.S., Silver, P.G., Liu, K.H. and K.S. Group (2002). Mantle discontinuities beneath southern Africa. *Geophysical Research Letters*, **29**, 129 1-4.
- Good, N. and DeWit, M.J. (1997). The Thabazimbi-Murchison Lineament of the Kaapvaal Craton, South Africa: 2700 Ma of episodic deformation. *Journal of the Geological Society of London*, **154**, 93-97.
- Grigne, C. and Labrosse, S. (2001). Effects of continents on Earth cooling: thermal blanketing and depletion in radioactive elements. *Geophysical Research Letters*, **8**, 2707-2710.
- Gripp, A.E. and Gordon, R.G. (1990). Current plate velocities relative to the hotspots incorporating the NUVEL-1 global plate motion model. *Geophysical Research Letters*, **17**, 1109-1112.
- Gripp, A.E., and Gordon, R.G. (2002). Young tracks of hotspots and current plate velocities. *Geophysical Journal International*, **150**, 321-361.
- Guillou, L., and Jaupart, C. (1995). On the effect of continents on mantle convection. *Journal of Geophysical Research*, **100**, p.24,217-24,238.
- Hanson, R.E. (2003). Proterozoic geochronology and tectonic evolution of southern Africa, *In*: M. Yoshida, B. F. Windley and S. Dasgupta (Editors), Proterozoic East Gondwana: supercontinent assembly and breakup. *Geological Society of London Special Publication*, **206**, 427-463.
- Hanson, R.E., Gose, W.A., Crowley, J.L., Ramezani, J., Bowring, S.A., Bullen, D.S., Hall, R.P., Pancake, J.A. and Mukwakwami, J. (2004). Geochronology, paleomagnetism and geochemistry of 1.9 Ga dolerite intrusions in the Waterberg Group, South Africa and Botswana. *South African Journal of Geology*, **107**, 233-254.
- Holt, W.E. (2000). Correlated crust and mantle strain fields in Tibet. *Geology*, **28**, 67-70.
- Horstwood, M., Nesbitt, R.W., Noble, S.R. and Wilson, J.F. (1999). U-Pb zircon evidence for an extensive early Archean craton in Zimbabwe: A reassessment of the timing of craton formation, stabilization, and growth. *Geology*, **27**, 707-710.
- Huerta, A.D., Royden, L. and Hodges, K. (1998). The thermal structure of collisional orogens as a response to accretion, erosion, and radiogenic heating. *Journal of Geophysical Research*, **103**, 15,287-15,302.
- James, D.E., Fouch, M.J., VanDecar, J.C., van der Lee, S. and K.S. Group (2001). Tectospheric structure beneath southern Africa. *Geophysical Research Letters*, **28**, 2,485-2,488.
- Jordan, T.H. (1975). The continental tectosphere. *Reviews of Geophysics and Space Physics*, **13**, 1-12.
- Jordan, T. H. (1988). Structure and formation of the continental tectosphere. *Journal of Petrol. Special Lithosphere Issue*, 11-37.
- Kamber, B.S., Blenkinsop, T.B., Villa, I.M. and Dahl, P.S. (1995). Proterozoic transpressive deformation in the Northern Marginal Zone, Limpopo Belt, Zimbabwe. *Journal of Geology*, **103**, 493-508.
- Karato, S.-I. and P. Wu (1993). Rheology of the upper mantle: A synthesis. *Science*, **260**, 771-778.
- Kreissig, K., Holzer, L., Frei, R., Villa, I. M., Kramers, J. D., Kröner, A., Smit, C. A. and van Reenen, D. D. (2001). Geochronology of the Hout River Shear Zone and the metamorphism in the Southern Marginal Zone of the Limpopo Belt, Southern Africa. *Precambrian Research*, **109**, 145-173.
- Kroner, A., Jaeckel, P. and Brandl, G. (2000). Single zircon ages for felsic to intermediate rocks from the Pietersburg and Giyani greenstone belts and bordering granitoid orthogneisses, northern Kaapvaal Craton, South Africa. *Journal of African Earth Sciences*, **30**, 773-793.
- Majaule, T., Hanson, R.E., Key, R.M., Singletary, S.J., Martin, M.W. and Bowring, S.A. (2001). The Magondi Belt in northeast Botswana: regional relations and new geochronological data from the Sua Pan area. *Journal of African Earth Sciences*, **32**, 257-267.
- Majaule, T., Hanson, R.E., Key, R.M., Singletary, S.J., Martin, M.W. and Bowring, S.A. (2001). The Magondi Belt in northeast Botswana: regional relations and new geochemical data from the Sua Pan area. *Journal of African Earth Sciences*, **32**, 257-267.
- Mapeo, R.B.M., Armstrong, R.A. and Kampunzu, A.B. (2001). SHRIMP U-Pb zircon geochronology of gneisses from the Gweta borehole, northeast Botswana: implications for the Palaeoproterozoic Magondi Belt in southern Africa. *Geological Magazine*, **138**, 299-308.
- McCourt, S. and Armstrong, R.A. (1998). SHRIMP U-Pb zircon geochronology of granites from the Central Zone, Limpoo Belt, southern Africa: implications for the age of the Limpopo Orogeny. *South African Journal of Geology*, **101**, 329-338.
- McCourt, S., Hilliard, P., Armstrong, R.A. and Munyanyiwa, H. (2001). SHRIMP U-Pb zircon geochronology of the Hurungwe granite northwest Zimbabwe: Age constraints on the timing of the Magondi orogeny and implications for the correlation between the Kheis and Magondi belts. *South African Journal of Geology*, **104**, 39-46.
- McNamara, D.E., Owens, T.J., Silver, P.G. and Wu, F.T. (1994). Shear-wave anisotropy beneath the Tibetan Plateau. *Journal of Geophysical Research*, **99**, 13655-13665.
- Mkweli, S., Kamber, B. and Berger, M. (1995). Westward continuation of the craton-Limpopo Belt tectonic break in Zimbabwe and new age constraints on the timing of the thrusting. *Journal of the Geological Society of London*, **152**, 77-83.
- Mukasa, S.B., Wilson, A.H. and Carlson, R.W. (1998). A multielement geochronologic study of the Great Dyke, Zimbabwe: significance of the robust and reset ages. *Earth and Planetary Science Letters*, **164**, 353-369.
- Nguuri, T., Gore, J., James, D.E., Webb, S.J., Wright, C., Zengeni, T.G., Gwavava, O., Snoke, J.A. and K. S. Group. (2001). Crustal structure beneath southern Africa and its implications for the formation and evolution of the Kaapvaal and Zimbabwe cratons. *Geophysical Research Letters*, **28**, 2501-2504.
- Niu, F. and James, D.E. (2002). Fine structure of the lowermost crust beneath the Kaapvaal craton and its implications for crustal formation and evolution. *Earth and Planetary Science Letters*, **200**, 121-130.
- Poujol, M., Robb, L.J., Respaut, J.-P. and Anhaeusser, C.R. (1996). 3.07-2.97 Ga greenstone belt formation in the northeastern Kaapvaal Craton: implications for the origin of the Witwatersrand Basin. *Economic Geology*, **91**, 1455-1461.
- Poujol, M., Anhaeusser, C.R. and Armstrong, R.A. (2002). Episodic granitoid emplacement in the Archean Amalia-Kraaipan terrane, South Africa: confirmation from single zircon U-Pb geochronology. *Journal of African Earth Sciences*, **35**, 147-161.
- Ramokate, L.V., Mapeo, R.B.M., Corfu, F. and Kampunzu, A.B. (2000). Proterozoic geology and regional correlation of the Ghanzi-Makunda area, western Botswana. *Journal of African Earth Sciences*, **30**, 453-466.
- Ribe, N.M. (1989). Seismic anisotropy and mantle flow. *Journal of Geophysical Research*, **94**, 4213-4223.
- Rümpker, G. and Silver, P.G. (1998). Apparent shear-wave splitting parameters in the presence of vertically-varying anisotropy. *Geophysical Journal International*, **135**, 790-800.
- Saltzer, R.L. (2002). Upper mantle structure of the Kaapvaal craton from surface wave analysis – a second look. *Geophysical Research Letters*, **29**, 17-1 through 17-4.
- Sandvol, E.A., Ni, J.F., Kind, R. and Zhao, W. (1997). Seismic anisotropy beneath the southern Himalayas-Tibet collision zone. *Journal of Geophysical Research*, **102**, 17,813-17,823.
- Schaller, M., Steiner, O., Studer, I., Holzer, L., Herwegh, M and Kramers, J.D. (1999). Exhumation of Limpopo Central Zone granulites and dextral continent-scale transcurrent movement at 2.0 Ga along the Palala Shear Zone, Northern Province, South Africa. *Precambrian Research*, **96**, 263-288.
- Schmitz, M. (2002). Geological and thermochronological studies of the evolution of the lower crust of southern Africa. *Unpublished Ph.D. thesis, Massachusetts Institute of Technology, Cambridge, Massachusetts, U.S.A.*, 1-271.

- Schmitz, M.D., Bowring, S.A., de Wit, M.J. and Gartz, V. (2004). Subduction and terrane collision stabilize the western Kaapvaal craton tectosphere 2.9 billion years ago. Submitted to *Earth and Planetary Science Letters*, in press.
- Sengor, A.M.C., Burke, K. and Dewey, J.F. (1978). Riffs at high angles to orogenic belts: Tests for their origin and the Upper Rhino Graben as an example. *American Journal of Science*, **278**, 24-40.
- Silver, P.G. and Chan, W.W. (1988). Implications for continental structure and evolution from seismic anisotropy. *Nature*, **335**, 34-39.
- Silver, P.G. and Savage, M. (1994). The interpretation of shear-wave splitting parameters in the presence of two anisotropic layers. *Geophysical Journal International*, **119**, 949-963.
- Silver, P.G. (1996). Seismic anisotropy beneath the continents: Probing the depths of geology. *Annual Review of Earth and Planetary Science*, **24**, 385-432.
- Silver, P.G., Mainprice, D., Ismail, W.B., Tommasi, A., and Barruol, G. (1999). Mantle structural geology from seismic anisotropy, *In*: Y Fei, C. Bertka and B. O. Mysen (Editors). *Mantle Petrology, Field Observations and High Pressure Experimentation: A Tribute to Francis R. (Joe) Boyd*. *Chemical Society, Special Publication*, **6**, 79-103.
- Silver, P.G., Gao, S.S., Liu, K.H. and K.S.Group. (2001). Mantle deformation beneath southern Africa. *Geophysical Research Letters*, **28**, 2493-2496.
- Silver, P.G. and Holt, W.E. (2002). The mantle flow field beneath western North America. *Science*, **295**, 1054-1057.
- Sleep, N.H. (2000). Evolution of the mode of convection within terrestrial planets. *Journal of Geophysical Research, Earth and Planets*, **105**, 17,563-17,578.
- Tinker, J., de Wit, M.J. and Grotzinger, J. (2002). Seismic stratigraphic constraints on Neoproterozoic evolution of the western margin of the Kaapvaal craton, South Africa. *South African Journal of Geology*, **105**, 107-134.
- Tommasi, A., and Vauchez, A. (2001). Continental rifting parallel to ancient collisional belts; an effect of the mechanical anisotropy of the lithospheric mantle. *Earth and Planetary Science Letters*, **185**, 199-210.
- Treloar, P.J. (1988). The geological evolution of the Magondi Mobile Belt, Zimbabwe. *Precambrian Research*, **38**, 55-73.
- Vinnik, L.P., Green, R.W.E., and Nicolaysen, L.O. (1995). Recent deformations of the deep continental root beneath southern Africa. *Nature*, **375**, 50-52.
- Walraven, F. and Hattingh, E. (1993). Geochronology of the Nebo Granite, Bushveld Complex. *South African Journal of Geology*, **96**, 31-41.
- White, R.S. and McKenzie, D. (1995) Mantle plumes and flood basalts. *Journal of Geophysical Research*, **100**, 17,543-17,585.
- Wilson, J. F. (1990). A craton and its cracks: some of the behaviour of the Zimbabwe block from the Late Archaean to the Mesozoic in response to horizontal movements, and the significance of some of its mafic dyke fracture patterns. *Journal of African Earth Sciences*, **10**, 483-501.
- Wingate, M.T. (2000). Ion microprobe U-Pb zircon and baddelyite ages for the Great Dyke and its satellite dykes, Zimbabwe. *South African Journal of Geology*, **103**, 74-80.

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